Retrieval of Model Initial Fields from Single-Doppler Observations of a Supercell Thunderstorm. Part II: Thermodynamic Retrieval and Numerical Prediction

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Submitted to *Monthly Weather Review*

22 February 2001

Revised

26 July 2001

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ABSTRACT

In this two-part study, a single-Doppler parameter retrieval technique is developed and applied to a real data case to provide model initial conditions for a short-range prediction of a supercell thunderstorm. The technique consists of the sequential application of a single-Doppler velocity retrieval (SDVR), followed by a variational velocity adjustment, a thermodynamic retrieval, and a moisture specification step. In Part I, we described the SDVR procedure and presented results from its application to a supercell thunderstorm. In Part II, we present results from the thermodynamic retrieval and the numerical model prediction for this same case. For comparison, we also show results from parallel sets of experiments using dual-Doppler derived winds and winds obtained from the simplified velocity retrieval described in Part I.

Following the SDVR, the retrieved wind fields (available only within the storm volume) are blended with a base-state background field obtained from a proximity sounding. The blended fields are then variationally adjusted to preserve anelastic mass conservation and the observed radial velocity. A Gal-Chen type thermodynamic retrieval procedure is then applied to the adjusted wind fields. For all experiments (full retrieval, simplified retrieval, and dual-Doppler), the resultant perturbation pressure and potential temperature fields agree qualitatively with expectations for a deep-convective storm. An analysis of the magnitude of the various terms in the vertical momentum equation for both the full retrieval and dual-Doppler experiments indicates a reasonable agreement with predictions from linear theory. In addition, the perturbation pressure and vorticity fields for both the full retrieval and dual-Doppler experiments are in reasonable agreement with linear theory predictions for deep-convection in sheared flow.

Following a simple moisture specification step, short-range numerical predictions are initiated for both retrieval experiments and the dual-Doppler experiment. In the full single-Doppler retrieval and dual-Doppler cases, the general storm evolution and deviant storm motion are reasonably well predicted for a period of about 35 minutes. In contrast, the storm initialized

using the simplified wind retrieval decays too rapidly, indicating that the additional information obtained by the full wind retrieval (primarily low-level polar vorticity) is vital to the success of the numerical prediction. Sensitivity experiments using the initial fields from the full retrieval indicate that the predicted storm evolution is strongly dependent on the initial wind and moisture fields. Overall, the numerical prediction results suggest at least some degree of short-term predictability for this storm and provide an impetus for continued development of single-Doppler retrieval procedures.

1. Introduction

This paper is the second in a two-part study describing a sequential single-Doppler parameter retrieval procedure designed to provide initial forecast fields for deep-convective storms. This procedure features the sequential application of a single-Doppler velocity retrieval (SDVR), a velocity blending and adjustment technique, a thermodynamic retrieval, and a simple moisture specification step (see Fig. 1 in Part I for a flow chart illustrating the entire procedure).

In Part I, we described the first component of the sequential single-Doppler retrieval, a reflectivity conservation-based SDVR (Shapiro et al. 1995a) and presented results from its application to a deep-convective storm. In Part II of this study, we describe the remainder of the sequential retrieval procedure, document its application to the three-dimensional vector wind fields obtained in Part I, and report on short-range numerical predictions initialized from the retrieved fields.

Obviously, the success of the numerical predictions will depend on more than just the quality of the retrieved fields. Two other key factors are the adequacy of the numerical prediction model and the predictability of the particular convective storm. With regard to the question of predictability, application of the Lorenz (1969) turbulence-based analysis suggests a predictability limit of about 2 h for features with a 20-km wavelength (Lilly 1990). For a single-cell thunderstorm with a life-cycle of 30-50 min, the predictability limit may be even shorter. Lilly notes, however, that the degree to which the Lorenz analysis can be applied to highly intermittent deep-convective phenomena is unclear, and more encouraging counter-examples can be found. These include rotating supercell storms, which appear to be more predictable than their nonrotating counterparts (Lilly 1986, Droegemeier et al. 1993), and strongly forced squall lines, which may to some degree “inherit” the predictability of the larger scale forcing mechanisms. Recognizing that a wide range of predictability limits likely exists for different convective phenomena, we have selected a fairly isolated supercell storm for this proof-of-concept study. This choice should maximize our chances for success, while simplifying our numerical predictions by allowing us to neglect the complexities of nonhomogeneous mesoscale forcing.

While the ultimate operational utility of explicit storm-scale numerical weather prediction will likely not be known for a number of years, our goal in Part II is to evaluate, for one particular case, the skill of an explicit storm-scale prediction initialized with single-Doppler retrieved fields. In section 2, the thermodynamic retrieval, velocity blending and adjustment procedure, and moisture specification step are described in detail. [See Part I for an overview of the various techniques available for diagnosing thermodynamic and microphysical fields from a time history of wind data.] In section 3, we describe the environmental conditions and evolution of the Arcadia, OK, supercell storm of 17 May 1981 (Dowell and Bluestein 1997) used in this study. In section 4, we present results from the application of the thermodynamic retrieval to the three-dimensional wind fields obtained using the mean wind moving reference frame in Part I. In section 5, we report on short-range numerical prediction experiments initialized with the dual-Doppler derived and single-Doppler retrieved fields. These experiments are conducted using the Advanced Regional Prediction System (ARPS) model (Xue et al. 2000, Xue et al. 2001, Xue et al. 1995). Finally, we summarize our results and discuss their implication for storm-scale prediction in section 6.

2. Retrieval procedure

a. Velocity blending and adjustment

The SDVR detailed in Part I yields the three spherical wind components on a spherical grid. These retrieved spherical components are then converted to Cartesian components and interpolated to the ARPS grid for use as input for the subsequent retrieval steps. Two practical limitations must still be overcome prior to application of the thermodynamic retrieval and initialization of the numerical prediction model. First, because the radar coverage is incomplete, the retrieved wind field does not cover the entire model domain and contains internal data "holes". Second, despite the use of mass conservation as a wind retrieval constraint, the retrieved wind field generally does not satisfy mass conservation on the model grid (due to discretization errors associated with the interpolation). Thus, it is desirable to "blend" the retrieved wind field with a background wind field and adjust the blended wind field to satisfy mass conservation on the model grid.

In this study, the available background field is limited to a vertical profile of mean horizontal wind components from a single proximity sounding. Accordingly, a very simple blending procedure (following Lin et al. 1993) is adapted with the goal of providing a smooth transition from the radar-retrieved horizontal wind vectors to the environmental winds (obtained from the proximity sounding). A series of two-dimensional horizontal Laplace equations are solved over the data void areas for each of the horizontal wind components. Dirichlet boundary conditions are specified along the edge of the retrieval area (obtained from the retrieved wind field) and along the lateral boundaries of the model domain (obtained from the proximity sounding). Thus, the retrieved wind vectors are retained within the storm and the blended vectors are used outside the storm. The vertical velocity is set equal to zero outside the storm (within the data-void region).

To enforce mass conservation over the entire model domain, a variational wind adjustment procedure described by Shapiro et al. (1995b) is applied to the blended wind field. Patterned after Liou's (1989) scheme, this procedure produces a wind field that minimizes departures from the input (blended) wind field, while simultaneously satisfying anelastic mass conservation, and matching the radar-observed radial velocity. An expression for the adjusted wind field is obtained by minimizing the following cost function:

$$J = \iiint_\sigma \Bigg[\left(\bar{\rho} u^{adj} - \bar{\rho} u^{bld}\right)^2 + \left(\bar{\rho} v^{adj} - \bar{\rho} v^{bld}\right)^2 + \left(\bar{\rho} w^{adj} - \bar{\rho} w^{bld}\right)^2$$

$$+ \lambda_1 \bar{\rho}\left(x u^{adj} + y v^{adj} + z w^{adj} - r v_r^{obs}\right) + \lambda_2 \left(\frac{\partial \bar{\rho} u^{adj}}{\partial x} + \frac{\partial \bar{\rho} v^{adj}}{\partial y} + \frac{\partial \bar{\rho} w^{adj}}{\partial z}\right)\Bigg] d\sigma = 0 \quad , \qquad (1)$$

where λ_1 and λ_2 are the Lagrange multipliers and σ represents the three-dimensional spatial domain. The superscript "bld" indicates the blended wind field obtained via the hole-filling procedure, and the superscripts "adj" and "obs" indicate the adjusted and observed velocities, respectively. Setting the variation of (1) with respect to each of the Cartesian velocity components equal to zero and integrating by parts yields the Euler-Lagrange equation

$$2\bar{\rho}\left(\vec{V}^{adj} - \vec{V}^{bld}\right) + \lambda_1 \vec{r} - \nabla\lambda_2 = 0 \qquad , \tag{2}$$

where $\vec{r}$ is the local position vector. On the lateral boundaries we set $\lambda_2 = 0$, while use of the $\delta w^{adj} = 0$ condition on the model top and bottom allows us to enforce the condition $w^{adj} = w^{bld}$, where w^{bld} is set equal to zero (the impermeability condition). Setting $w^{adj} = w^{bld}$ in the vertical component of (2) yields the λ_2 boundary condition

Each of the two strong constraints in (1) can be used to eliminate $\vec{V}^{adj}$ in (2), yielding a coupled system of equations for λ_1 and λ_2 involving known quantities:

$$\lambda_1 = \frac{2}{r^2}\left[\bar{\rho} r v_r^{bld} - \bar{\rho} r v_r^{obs} + \frac{1}{2}\vec{r}\cdot\nabla\lambda_2\right] \tag{3}$$

$$\nabla^2\lambda_2 = -2\nabla\cdot\bar{\rho}\vec{V}^{bld} + \nabla\cdot\left(\vec{r}\lambda_1\right) \quad . \tag{4}$$

Eliminating λ_1 between (3) and (4) results in a two-dimensional elliptic equation for λ_2 on spherical surfaces:

$$\nabla^2_{\theta,\phi}\lambda_2 = -2\nabla\cdot\bar{\rho}\vec{V}^{bld} - \frac{2}{r^2}\frac{\partial}{\partial r}\left[r^2\bar{\rho}\left(v_r^{obs} - v_r^{bld}\right)\right] \quad . \tag{5}$$

An explicit form of the top and bottom boundary condition is obtained by setting $w^{adj} = w^{bld}$ in the vertical component of (2) and using (3) to eliminate λ_1:

$$\frac{\partial\lambda_2}{\partial z} = \frac{z}{x^2+y^2}\left[2\bar{\rho} r\left(v_r^{bld} - v_r^{obs}\right) + x\frac{\partial\lambda_2}{\partial x} + y\frac{\partial\lambda_2}{\partial y}\right] \quad . \tag{6}$$

Shapiro et al. (1995b) expressed (6) in Cartesian coordinates so that it could be solved on the ARPS grid. Attemps to solve the Cartesian formulation via a successive-under-relaxation technique resulted in very slow convergence. We therefore elected to iteratively solve (4) and (5) for λ_1 and λ_2 on the ARPS model grid, using a successive-under-relaxation technique for λ_2 (starting with a first guess of zero for λ_1). This formulation has two known deficiencies related to the use of radial velocity as a strong constraint. First, blended values for radial velocity must be used outside the region of radar observations. Second, a region of singularity exists at and above the radar, where the horizontal projection of $\vec{r}$ goes to zero.

b. Thermodynamic retrieval and moisture specification

Once the retrieved wind field has been blended with the background wind field and variationally adjusted, the resultant fields are used as input for the thermodynamic retrieval. The formulation follows Gal-Chen's (1978) procedure in which the horizontal momentum equations are used as weak constraints to compute the two-dimensional horizontal pressure field from known forcing terms. The complete three-dimensional pressure field is obtained by solving a series of 2-D Poisson equations on horizontal model levels, using the Dirichlet condition of $p' = 0$ along the lateral boundaries. This is appropriate for our simple case, where a proximity sounding has been used to specify background horizontal winds at lateral boundaries that are well displaced from the storm. We have also successfully tested the use of Dirichlet conditions for real data cases where the perturbation pressure along the lateral boundary is obtained from a full three-dimensional analysis of pressure (Shapiro et al. 1996, Weygandt et al. 1999b). Neumann boundary conditions, however, should be used for any portion of the lateral boundary that is crossed by the storm volume.

Within the radar reflectivity region, the forcing function for the pressure retrieval is calculated on the ARPS model grid, using model routines. Following Ellis (1997), the forcing function is set to zero outside the reflectivity region and the pressure retrieval reduces to a 2-D Laplace equation for perturbation pressure.

Once a unique perturbation pressure field is obtained, the vertical momentum equation can be solved for the perturbation potential temperature, provided the distribution of moisture variables is known:

$$\theta' = \frac{\bar{\theta}}{\bar{\rho} g}\frac{\partial p'}{\partial z} + \frac{\bar{\theta}}{\bar{\rho} g}\left[\frac{\partial(\bar{\rho} w)}{\partial t} + \vec{V}\cdot\nabla(\bar{\rho} w)\right] + \frac{\bar{\theta}}{\bar{\rho} g}\frac{p'}{\gamma \bar{p}} - .608\bar{\theta}\, q_v{}' + \bar{\theta}\, q_c + \bar{\theta}\, q_r + \frac{\bar{\theta}}{\bar{\rho} g} F_z \quad . \qquad (7)$$

In (7), q_v, q_c, and q_r are mixing ratios for water vapor, cloud water, and rainwater, the three moisture variables in the Kessler (1969) microphysical parameterization used in this study. Rainwater mixing ratios are computed from the observed radar reflectivity using a semi-empirical relationship from Kessler (1969). Since we are not accounting for ice physics in the retrieval or subsequent numerical predictions, the radar-observed reflectivity is truncated at 50 dBZ prior to the calculation of the rainwater mixing ratio. Water vapor and cloud water effects are neglected in the thermodynamic retrieval. Again, following Ellis (1997), and based on our own experience, the retrieved pressure and potential temperature are only retained within the reflectivity volume. Outside the reflectivity region, pressure and temperature are set to base-state values (obtained from the proximity sounding).

As a last step in the retrieval process, the water vapor mixing ratio is modified within the radar-observed reflectivity region. Specifically, regions where the rainwater exceeds 0.1 g kg^{-1} and the retrieved vertical velocity exceeds 3 m s^{-1} are saturated using the retrieved potential temperature. The choices for the rainwater mixing ratio and vertical velocity thresholds, while obviously somewhat arbitrary and grid-resolution-dependent, are designed to only saturate strong updrafts contained within the storm volume.

The saturated water vapor mixing ratio is obtained with Teten's formula and no attempt is made to specify the cloud water field. Based on the results of Weygandt et al. (1999a), however, effects from this omission should be minimal, as the cloud water field typically develops quite rapidly from the other moisture fields. Of likely greater significance is the neglect of frozen hydrometeors in both the specification of the initial fields and the subsequent numerical prediction. Our strategy here is to examine the retrieval and prediction performance using the relatively simple warm rain parameterization, recognizing that the difficult ice-phase microphysical retrieval problem remains.

3. The 17 May 1981 supercell case

In Part I, we described the dual-Doppler dataset from the 17 May 1981 Arcadia, OK tornadic supercell case and its processing. Here, a brief overview of the environmental conditions and storm evolution is presented as an aid for evaluating the numerical prediction results presented in section 5. Refer to Dowell and Bluestein 1997 (hereafter DB97) for a more detailed description of the case. Meteorological conditions on 17 May 1981 featured a closed circulation 500-mb short-wave trough moving from the southern Rockies into the southern Plains. By 2100 UTC, an associated surface cyclone was located along the Kansas-Oklahoma border north of Enid, OK, with a warm front extending eastward and a dryline extending southward. Figure 1a shows the 2115 UTC special sounding obtained at Tuttle, OK (smoothed and interpolated to a uniform vertical grid). The thermodynamic profile shows warm, moist air near the surface, with a pronounced dry layer between 700 and 500 mb. Much of the troposphere was conditionally unstable, with convective available potential energy (CAPE) in excess of 3000 J kg^{-1}. Also evident in the sounding were strong southwesterly winds at mid and upper levels of the troposphere. The environmental hodograph (Fig. 1b) shows 0-4 km shear to be in excess of $5(10)^{-3}$ s^{-1}.

Thunderstorms initially developed around 2000 UTC near the warm front dryline intersection in northcentral Oklahoma. Two isolated supercells subsequently developed in central Oklahoma just before 2100 UTC. The first supercell formed near Pocasset, OK, and moved northeastward through the National Severe Storms Laboratory dual-Doppler network, producing an F2 tornado south of Arcadia, OK. That storm is the focus of this study and hereafter is referred to as the Arcadia supercell. DB97 have performed a detailed dual-Doppler study of this storm and made their analyses available to be used as input for comparison thermodynamic retrievals and model initializations. The second storm formed near Rush Springs, OK, and also moved northeastward, producing an F3 tornado near Tecumseh, OK, and an F4 tornado near Okemah, OK (Brewster 1984).

Figure 2 depicts the evolution of the Arcadia supercell reflectivity echo from 2239 UTC (the middle time of the retrievals and the initialization time of the numerical predictions) through the 10-min tornadic phase (centered at 2305 UTC), to the posttornadic weakening phase at 2322 UTC. Individual hydrometeor cores are identified and labeled at each time. These include the main core (M) and a core to its east (E), two left-moving cores (L1, L2), and a core that develops to the southeast of the main core as the storm weakens (SE). In section 5, these cores and low-level cyclonic circulation centers associated with them will be used to evaluate the fidelity of the prediction experiments. At 2239 UTC, a left-moving hydrometeor core (L1) can be seen to the north of the main core (M), which by 2251 UTC was evolving toward its tornadic phase and beginning to shed another left-moving core (L2). The tornado is near maximum intensity at 2305 UTC and has dissipated by 2313 UTC, at which time the L2 core has become nearly detached from the M core. Also at 2313 UTC, a new core is beginning to form to the southeast of the M core (SE). By 2322 UTC, the L2 core has moved northward away from the weakening M core, as the new SE core intensifies.

4. Retrieval results

We now present results from the application of the velocity blending and adjustment, thermodynamic retrieval, and moisture specification to the wind fields retrieved using the mean wind moving reference frame in Part I (denoted SDVR). For comparison we also show results for the corresponding dual-Doppler wind fields (denoted DDOP) and discuss results for the drastically simplified retrieval described in Part I (denoted UVVR). The simplified retrieval (UVVR) retains only the observed radial velocity, estimated mean horizontal wind components, and perturbation radial convergence contribution to the polar velocity (obtained from mass conservation). These three wind field datasets, which are also used for the numerical prediction experiments presented in section 5, are summarized in Table 1.

a. Blended and adjusted wind field

Figure 3 shows the z = 2.25 km hole-filled perturbation horizontal wind vectors for the DDOP and SDVR cases. For both cases, the perturbations are computed as deviations from the horizontally homogeneous wind field obtained from the proximity sounding shown in Fig. 1. Consistent with the specification of Dirichlet lateral boundary conditions, both vector wind fields exhibit a smooth transition from the radar-derived values along the interior (storm) boundary to near zero along the exterior (domain) boundary (note that the storm extends beyond the 0.1 g kg^{-1} contour). The most significant difference between the SDVR and DDOP fields within the blended region (outside the storm) is the area of strong west-northwesterly winds between the primary storm and the storm in the northwest portion of the domain. This difference is because the DDOP dataset (analyzed independently by DB97) retained no wind information from the smaller northern storm, resulting in a smooth transition from the northern edge of the main storm to the domain boundary. In contrast, for the SDVR case, the hole-filling procedure produced a smooth transition between the retrieved westerly winds in both storms.

A vertical cross section through the updraft/mesocyclone of the storm (the N-S line at x = 19.5 km in Fig. 3) is shown in Fig. 4. While both the DDOP and SDVR cases show the strong storm updraft, significant differences are evident between the SDVR and DDOP fields. Most noticeably, the SDVR case is missing the strong vertical velocities and divergence that are evident near the top of the storm in the DDOP case. This difference is due in part to the reduced retrieval coverage area associated with the mean wind moving reference frame. Because the speed of the storm was slower than the mean wind, the mean wind moving reference frame "overshifted" the first and third time-levels of data. The cross-section of retrieved vectors from the UVVR case (not shown) closely resembles the SDVR cross-section.

Once the horizontal velocity fields have been hole-filled and the vertical velocity set equal to zero outside the reflectivity region, the variational velocity adjustment procedure is applied to enforce anelastic mass conservation, while preserving the observed radial velocity. Solution convergence for the successive underrelaxation solver (relaxation coefficient = 0.25) is determined via a grid-length-dependent iteration threshold for λ_2. Converged solutions are obtained after about 450 iterations for both the SDVR and DDOP cases and yield about a decadal decrease in the maximum anelastic divergence at each level. Figure 5 shows the vertical profile of maximum three-dimensional divergence before and after the adjustment for the SDVR case.

The impact of the variational velocity adjustment on the SDVR maximum vertical velocity profile is shown in Fig. 6. The profile has been modified to better fit the classic "bowstring" shape. This adjustment, similar to that produced by an O'Brien (1970) correction, is consistent with the enforcement of impermeable top and bottom boundary conditions in the mass conserving adjustment. For the DDOP case, in which the maximum vertical velocity profile already has a classic bowstring shape (see Fig. 5 in Part I), the profile is only slightly modified by the adjustment procedure.

b. Retrieved pressure fields

For the SDVR case, the retrieved perturbation pressure field (Fig. 7a) shows a pressure couplet associated with the southernmost updraft, and pressure signatures associated with the northern two updrafts. In all three cases, the pressure features are aligned with the shear vector through the updraft in a manner consistent with the linear theory of Rotunno and Klemp (1982). This theory predicts the existence of low (high) pressure perturbations down- (up-) shear from the updraft due to the interaction of the updraft with the environmental shear. Figure 7b shows that each updraft is also associated with a vertical vorticity couplet aligned perpendicularly to the shear vector, again in qualitative agreement with the linear theory of Rotunno and Klemp (1982) and Davies-Jones (1984). Pressure and vorticity fields for the DDOP case (not shown) are noisier, but show similar patterns.

N-S vertical cross sections (x = 19.5) of the retrieved pressure field for the DDOP and SDVR cases are shown in Fig. 8. The fields are qualitatively similar, with the DDOP case exhibiting smaller-scale features and larger amplitudes. The cross section of retrieved pressure for the UVVR case (not shown) looks quite similar to that for the SDVR case. At upper levels, all three cases exhibit a N-S couplet of positive and negative perturbation pressure, with the positive region extending downward into the central portion of the storm. Outside the storm region, where the forcing function is set equal to zero, the retrieved perturbation pressure smoothly transitions to near zero values along the domain edge. The retrieved pressure and potential temperature from this region are still discarded prior to the model initialization.

Calculations of the momentum checking measure of retrieval error (E_r , defined by Gal-Chen 1978) over the three-dimensional retrieval volume yields values of 0.52 for the DDOP and SDVR cases compared with 0.42 for the UVVR case. These values are quite large, indicating a rather poor fit of the horizontal pressure gradients to the forcing functions. As noted by Sun and Crook (1996), however, the momentum checking measure of retrieval error should be used with caution, because it does not respond to rotational versus divergent forcing errors in the same manner that the retrieved pressure field responds to these errors. A purely rotational error can be added to the forcing function without altering the retrieved pressure.

c. Potential temperature and moisture fields

Fig. 9a shows a N-S vertical cross section (x = 19.5 km) of the retrieved perturbation potential temperature over a portion of the domain for the SDVR cases. The field exhibits a narrow column of positive perturbation potential temperature that coincides quite well with the main storm updraft (see Fig. 4b). The maximum perturbation potential temperature of nearly +7 K (near z = 8 km) is in reasonable agreement with a parcel-theory-based estimate of +9 K obtained from the proximity sounding (see Fig. 1a). To the right (north) of the main updraft, an area of negative perturbation temperature is associated with the downdraft depicted in Fig. 4b.

The retrieved perturbation potential temperature for the UVVR experiment (not shown) is nearly identical to that of the SDVR experiment. The DDOP retrieved perturbation potential temperature (not shown) is qualitatively similar to the SDVR field, but considerably noisier. One major difference of the DDOP field is an area of strongly negative perturbation potential temperatures near the storm top that is consistent with the inferred equilibrium level from the proximity sounding. At lower levels, all three cases appear to capture the warm inflow into the southern side of the storm. Conspicuously absent from any of the cases, however, is a low-level cold pool. Attempts to retrieve the low-level cold pool were thwarted by extreme sensitivity to the lower boundary condition when radar data were extrapolated to the ground. Consequently, we abandoned the downward extrapolation of the radar data and all attempts to retrieve the cold pool. Numerical prediction sensitivity experiments presented for this case in section 5 and for an idealized supercell in Weygandt et al. (1999a) indicate that the low-level cold pool generates very quickly from the rainwater field.

Figures 9b-d show the contributions to the retrieved perturbation potential temperature from the three leading terms in equation (7): vertical acceleration, rainwater loading, and vertical pressure gradient. The vertical acceleration term (shown in Fig. 9b) exhibits a narrow column of positive perturbation near the updraft location (y = 25 km) with a maximum in excess of + 2K.

In contrast, the vertical pressure gradient term is more horizontally stratified, but also contributes significantly (more than + 2 K) to the perturbation potential temperature maximum in the updraft. Also clearly evident in both of these terms are the contributions to the negative perturbation potential temperature in the downdraft region to the north of the main updraft. The corresponding DDOP contributions (not shown) are again qualitatively similar, but substantially noisier, and appear to contain a larger contribution from the vertical acceleration term.

To further examine the plausibility of the magnitudes of the various contributions to the retrieved perturbation potential temperature, we now compare the ratio of the vertical acceleration to the buoyancy for the retrieved fields with estimates of the value of the ratio obtained from linear theory. A simple expression for this ratio, which also expresses the degree of deviation from hydrostatic balance, can be easily derived. Consider the instantaneous vertical acceleration produced by buoyancy in a two-dimensional Boussinesq fluid at rest. The governing equations for the linear system are:

$$\frac{\partial w}{\partial t} + \frac{\partial P}{\partial z} = B \quad , \tag{8}$$

$$\frac{\partial u}{\partial t} + \frac{\partial P}{\partial x} = 0 \quad , \tag{9}$$

$$\frac{\partial u}{\partial x} + \frac{\partial w}{\partial z} = 0 \quad , \tag{10}$$

where $P = p'/\rho_0$ and $B = -g\rho'/\rho_0$. Equations (8) and (9) can be combined using (10) to obtain a Poisson equation for P:

$$\frac{\partial^2 P}{\partial x^2} + \frac{\partial^2 P}{\partial z^2} = \frac{\partial B}{\partial z} \quad . \tag{11}$$

If a cellular initial buoyancy field is specified:

$$B = b \sin\left(\frac{\pi}{L} x\right) \sin\left(\frac{\pi}{H} z\right), \tag{12}$$

where L and H are the horizontal and vertical wavelengths, respectively, then (11) can be solved for P:

$$P = \frac{HL^2}{\pi\left(H^2 + L^2\right)} b \sin\left(\frac{\pi}{L}x\right)\cos\left(\frac{\pi}{H}z\right) \quad . \tag{13}$$

The vertical derivative of P can then be calculated and substituted into (8) along with (12) to obtain the following relationship between the initial vertical acceleration and the initial buoyancy:

$$\frac{\partial w}{\partial t} = \left(\frac{H^2}{H^2 + L^2}\right) b \quad . \tag{14}$$

This simple formula expresses the ratio of the vertical acceleration to the buoyancy as a function of the aspect ratio $\left(H/L\right)$ of the initial buoyancy field, and has the following physical interpretation [as discussed by Houze (1993) and others]. The initial buoyancy perturbation induces an opposing vertical pressure gradient force [termed the buoyancy pressure gradient force by Houze (1993)] that partially offsets the vertical acceleration produced by the buoyancy field. The degree to which this adverse pressure gradient offsets the buoyancy depends on the aspect ratio of the buoyancy and provides a measure of the degree to which the flow is hydrostatic.

The ratio of the maximum vertical acceleration to the maximum buoyancy can be easily estimated for the various retrievals. Choosing a location near the approximately collocated vertical acceleration and buoyancy maxima, subjective estimates of 0.57 and 0.67 have been obtained for the SDVR and DDOP ratios, respectively. Determining the aspect ratio (H/L) of the buoyancy field is a bit more difficult; however, subjective estimates of 1.3 and 1.8 have been obtained for the SDVR and DDOP cases, respectively. Using these values, Fig. 10 shows the ratio of the vertical acceleration to the buoyancy plotted against the aspect ratio of the buoyancy field. Comparisons of both the SDVR and DDOP points with the linear theory curve indicate good agreement. As expected, the dual-Doppler case is less hydrostatic, possessing a taller, narrower buoyancy field and a larger vertical acceleration.

As discussed earlier, the radar-observed reflectivity is converted to rainwater mixing ratio prior to the thermodynamic retrieval. The final step in the preparation of the initial forecast fields is the water vapor specification, in which the updrafts within the radar reflectivity region are saturated. A N-S cross section (x = 19.5 km) of the resultant perturbation water vapor field is shown in Fig. 11. As can be seen, the maximum perturbation is about 7.5 g kg^{-1} near the bottom of the pronounced dry layer, just above the capping inversion (see Fig. 1a).

5. Numerical prediction results

The complete set of retrieved fields ($u, v, w, p', \theta', q_v', qr$) for each of the three wind field datasets (summarized in Table 1) are used to initialize 43-min high-resolution numerical predictions of the supercell evolution. We then evaluate the relative skill of the single-Doppler velocity retrieval prediction (SDVR) against both the dual-Doppler prediction (DDOP) and the prediction using the simple wind retrieval (UVVR). It is important to again emphasize that the success of the predictions is a function not only of the quality of the initial fields, but also of the adequacy of the numerical model and of the predictability of this particular deep-convective storm. We supplement the single- and dual-Doppler predictions by examining the impact of the initial water vapor and rainwater fields on the predicted storm evolution in a set of sensitivity experiments.

Before presenting our prediction results, it is instructive to consider the skill of simpler initialization techniques for this particular case. This will provide us with a benchmark for evaluating our prediction results and help us assess whether our initialization technique provides skill over simpler methods. One common method for initializing storm-scale models is to use a thermal perturbation superposed upon horizontally homogeneous background fields to initiate convection (Schlesinger 1975, Klemp and Wilhelmson 1978, Weisman and Klemp 1984, Brooks et al. 1993, Janish et al. 1995, Wicker et al. 1997, and many others). When used with observed profiles of wind, temperature and moisture, this technique has in some instances reproduced observed storms with a high degree of realism (Wilhelmson and Klemp 1981, Klemp et al. 1981). A detailed description of a "bubble" simulation, completed using the proximity sounding shown in Fig. 1, is presented in the Appendix. This idealized simulation produces a storm that bears some resemblance to the observed Arcadia storm; however, the predicted storm begins weakening after 40 min. In contrast, the observed Arcadia storm persists for more than 3 h. Thus, while the idealized simulation suggests the possibility for storms to briefly exhibit supercellular characteristics, it significantly underestimates the longevity of the observed storm.

a. Numerical model formulation

For this study, we use the ARPS (Xue et al. 2000, Xue et al. 2001, Xue et al. 1995), a three-dimensional nonhydrostatic, compressible model. The model equations are solved on an Arakawa C grid, using a mode-splitting technique in which the acoustic terms are calculated on a smaller time step than that used for the other terms. A second-order centered scheme with an Asselin (1972) time filter is used for the larger time step, and a first-order forward-backward scheme is used for the smaller time step. Spatial differencing is accomplished using a second-order quadratically conserving scheme.

For this application, a Kessler (1969) explicit warm-rain microphysics scheme is used, whereby conservation equations for water vapor, cloud water, and rainwater mixing ratios are solved. Subgrid-scale turbulence is parameterized using a diagnostic first-order closure (Lilly 1962, Smagarinsky 1963). A wave-radiation open boundary condition is used for the lateral boundaries (Klemp and Lilly 1978, Durran and Klemp 1983), while rigid boundaries are used for the model top and bottom. The prediction experiments are completed using a 100 x 100 x 17 km fixed domain with horizontal and vertical grid resolutions of 1000 m and 500 m, respectively.

b. Single- and dual-Doppler prediction results

Our assessment of the numerical prediction results includes both qualitative and quantitative evaluations. First, qualitative comparisons of the low-level ($z = 2.25$ km) rainwater and horizontal wind vectors are made with the corresponding dual-Doppler analyzed verification fields. In these comparisons, individual hydrometeor cores and their associated low-level cyclonic circulation centers are identified and tracked throughout the predicted and actual storm evolution. The relative positions of these features are then compared against storm motion projections from the 0-6 km mean wind and from persistence of the initial storm motion. Second, the predictions are quantitatively verified by computing correlation coefficients between the predicted rainwater fields and rainwater fields derived from the radar-observed storm. Time series of predicted and verifying domain maximum vertical velocity and vertical profiles of horizontal vertical vorticity are also compared.

1) Evolution of low-level fields

Figures 12a-d show the z = 2.25 km rainwater mixing ratio and horizontal vector fields for the 2251 UTC dual-Doppler verification and the three corresponding 12-min predictions. Indicated on each of the plots are the main hydrometeor core (M) and a weaker eastern core (E), as well as the two-dimensional rainwater correlation coefficients for each of the three prediction experiments. Note that the extremely large rainwater values in the main core of the verification analysis (Fig. 15a) are due to the extreme sensitivity of the semi-empirical $Z - q_r$ relationship, [eq. (19)], for values of reflectivity in excess of 60 dBZ, which are likely dominated by hail. At 2251 UTC, the Arcadia supercell was producing golf-ball sized hail (Taylor 1982) and had a reflectivity maximum of about 65 dBZ. Although all three experiments have similarly high rainwater correlation coefficients, the main hydrometeor core in the UVVR case is significantly weaker than the other two experiments and lacks the low-level mesocyclone evident in the DDOP and SDVR predictions.

Despite the obvious similarities between the verification and the DDOP and SDVR predictions, significant wind field differences exist, especially in the area north of the main hydrometeor core. Also, a number of spurious weak hydrometeor cores have developed around the DDOP storm. Of these, the core to south of the main storm (x = 30, y = 20) is the most significant, as will be seen at the next verification time. Hereafter, it (and a corresponding feature that develops in the SDVR prediction) is referred to as the southern hydrometeor core (S).

Figures 13a-d show the verification fields at 2305 UTC, as well as the corresponding 26-min predictions. The observed storm was producing an F2 tornado at this time, and as Fig. 13a shows, the well-defined low-level vortex signature is south of the main core. In the DDOP prediction, the main core is displaced only slightly to the north of the verification core, but the southwestern portion of the storm has weakened too rapidly. Nearly in its place is the rapidly developing southern hydrometeor core, which is beginning to merge with the main core. The SDVR prediction shows a very similar evolution, with the main core weakening too rapidly and being replaced by a new core forming just to its south. The UVVR prediction continues to be too weak and the storm appears to be dissipating. Thus we see that although the azimuthal component of the polar vorticity (which was the principal addition in the SDVR experiment compared to the UVVR experiment as explained in Part I) has little impact on the thermodynamic retrieval, it has a strong positive impact on the numerical forecast.

It is not entirely clear why the main hydrometeor cores weaken too rapidly and new cells form to the south in the SDVR and DDOP cases. Detailed examination of the near-surface fields (not shown), however, indicates that the gust front surges too rapidly toward the southeast, causing the main updraft core to become separated from the warm, moist boundary layer inflow. The predicted evolution of the rainwater, cold pool and gust front is somewhat problematic. On the one hand, the model is initialized without a low-level cold pool at a point in the storm evolution when one likely existed. On the other hand, the use of a warm-rain microphysics scheme as opposed to a more realistic ice-phase scheme favors heavier, more localized precipitation, a stronger cold pool, and a more rapidly surging gust front (Johnson et al. 1993). While the combined impact of these two known deficiencies is unclear, the model predictions rapidly generate a cold pool, and by 2305 UTC the model-predicted gust front is too strong. Sensitivity experiments that illustrate the strong influence of the initial rainwater field on the formation of the low-level cold pool are presented in section 5c and in Weygandt et al. (1999a).

The rainwater and horizontal wind fields at 2313 UTC, 3 min after the demise of the tornado and 34 min into the numerical prediction, are shown in Figs. 14a-d. Overall, they indicate increasing disagreement between the model prediction and the verification as indicated by the reduced two-dimensional correlation coefficients. In addition, all three prediction experiments lack the large, weak rainwater region surrounding the main storm core. This is likely due to the lack of ice species in the microphysical scheme and associated inability of the predicted storm to generate realistic anvils (McCumber et al. 1991, Johnson et al. 1993). Note, however, that in both the DDOP and SDVR predictions, the southern hydrometeor core corresponds quite favorably with the main core from the verification. In particular, the location of the DDOP southern core and the associated cyclonic flow pattern match that of the verification very closely. None of the predictions reproduce the developing southeastern core seen in the verification. The predicted fields from the final verification time (2322 UTC, 43 min-forecast, not shown), indicate that most of the predictive skill has been lost by all the experiments, as the observed storm transforms into a weakening multicellular storm.

The overall accuracy of the predicted storm movement is evaluated by comparing the locations of the z = 3.25 km cyclonic circulation centers at various times. Use of this particular feature proved to be a precise method for tracking motion of the hydrometeor cores, as a coherent circulation center was identified for each of the labeled hydrometeor cores in Figs. 12-14. Figure 15 shows the location of these centers at the model initial time (2239 UTC) and four subsequent verification times (2251, 2305, 2313, and 2322 UTC). The main circulation centers for all three prediction experiments are shown in Fig. 15a, and the DDOP and SDVR southern circulation centers are shown in Fig. 15b (UVVR never developed a southern core or circulation center). The location of the observed main circulation center is shown on both plots, as are extrapolated circulation centers based on the 0-6 km mean wind and persistence of the initial circulation center motion.

Figure 15a indicates that all three experiments predict the location of the main circulation center quite accurately at 2251 UTC, but then the predicted circulation centers begin to move with the mean wind as they prematurely weaken. Figure 15b shows that the DDOP southern circulation center (which developed at 2251 UTC) initially moves with the mean wind, but after 2305 UTC begins to deviate to the right of the mean wind. The SDVR southern circulation develops around 2305 UTC and only deviates slightly from the mean wind. Despite the discrepancies in the predicted storm evolution for the DDOP and SDVR cases, both of these experiments place a circulation center in much better agreement with the 2313 UTC observed main circulation center than the UVVR experiment or either of the two extrapolation techniques.

2) Quantitative verification scores

To better quantify the accuracy of the model-predicted storms, we now present some simple verification scores. We begin with the three-dimensional rainwater correlation between each of the three prediction experiments and the verification (Fig. 16). It is important to note that the computed correlation coefficients are strongly dependent on the overall size of the numerical domain and the formula used for converting the radar-observed reflectivity to rainwater. As such, the correlation scores should only be used to evaluate the relative skill of these experiments for this case. Two factors account for the nonunity correlation at the initial time. First, reflectivity observed by the Norman radar is converted to rainwater for the verification, whereas reflectivity from Cimarron is used to compute the initial rainwater for the prediction experiments. Second, as discussed in section 2b, reflectivity is truncated at 50 dBZ before conversion to rain water for the initial forecast fields.

For all three cases, the rainwater correlation coefficient decreases at a fairly steady rate throughout the 43-min prediction period. After a brief period of rapid decrease for the DDOP prediction, the DDOP and SDVR predictions have nearly identical correlation scores through 26 min. In contrast, the UUVR correlation coefficient is significantly lower than the others by 12 min and remains that way until the very end of the prediction, when all scores are quite poor. This result confirms that the addition of the low-level polar vorticity (obtained from the retrieved azimuthal velocity) in the full single-Doppler velocity retrieval (SDVR) significantly improves the numerical prediction compared to that for the simplified wind retrieval (UVVR). Furthermore, the similarity between the SDVR and DDOP scores suggests that the full wind retrieval captures those flow features contained in the dual-Doppler analysis that are most important to the storm evolution.

Last, we note that the three-dimensional correlation coefficients (Fig. 16) are much lower than the $z = 2.25$ km two-dimensional correlation coefficients presented with Figs. 12-14. Examination of vertical profiles of two-dimensional correlation coefficients (not shown) indicates consistently high correlation values through the lowest 8 km, with substantially lower correlations above that level. This dichotomous behavior seems to be related to rapid decrease in the depth of the predicted storms and indicates that the three-dimensional correlations may give an overly pessimistic estimate of the accuracy of the predicted rainwater at low- and midlevels.

We next examine model-predicted and dual-Doppler-derived domain maximum vertical velocity time series (Fig. 17). The dual-Doppler verification maximum (obtained from the independent analysis of DB97) is in excess of 50 m s^{-1} before the 14-min data gap and near 30 m s^{-1} after the data gap. Following a brief surge, the DDOP maximum rapidly decreases to values near 30 m^{-1} s^{-1}. Starting from a value near 20 m s^{-1}, the SDVR maximum rapidly increases to a similar value. Thereafter, the differences between the DDOP and SDVR maxima appear to be related to the timing of the development of their respective southern cores. In contrast, the maximum vertical velocity for the UVVR prediction stays near 20 m^{-1} s^{-1} for the first 20 min, then steadily decreases through the remainder of the prediction.

As a final verification measure, the time evolution of the vertical profiles of maximum vertical vorticity are examined for the DDOP and SDVR cases. These profiles (not shown) indicate that both experiments intensify the midlevel mesocyclone too rapidly while failing to reproduce the intensification of the low-level mesocyclone. A detailed search for the cause of this shortcoming is beyond the scope of this paper; however, the excessive surge of the surface gust front appears to play a role.

c. Moisture sensitivity experiments

We now present results from a set of three moisture sensitivity experiments (summarized in Table 2) that illustrate the importance of the initial water vapor and rainwater fields on the predicted storm evolution. Because single-Doppler retrieval is the focus of this study (and noting that the SDVR prediction performs nearly as well as the DDOP prediction), we chose the SDVR prediction as the control case. In the first experiment (NOQV) we specify the initial perturbation water vapor to be zero over the entire domain (base-state water vapor is retained). In the second experiment (NOQR) the initial rainwater is set equal to zero over the entire domain (and the retrieved potential temperature neglects the rainwater contribution). In the third experiment (NOQQ) both the perturbation water vapor and rainwater (and rainwater contribution to potential temperature) are set to zero. Analysis of the sensitivity experiments is accomplished by comparing time series of domain maximum vertical velocity and three-dimensional rainwater correlation, as well as qualitative comparison of selected fields.

Figure 18 shows the time series of domain maximum vertical velocity for the SDVR control and the three moisture sensitivity experiments. Both of the predictions in which the perturbation water vapor is withheld (NOQV and NOQQ) experience a rapid decrease in the domain maximum vertical velocity. This is consistent with the OSSE experiment results of Weygandt et al. (1990) and is easily understood by noting that unsaturated upward motion in a conditionally unstable atmosphere leads to adiabatic cooling of the parcel, thereby reducing buoyancy and upward acceleration. The upward motion and associated cooling, however, also lead to a rapid resaturation of the updraft, and the maximum vertical velocity quickly recovers for the NOQV experiment. In contrast, when neither the perturbation water vapor nor the rainwater are present in the initial fields (NOQQ), the maximum vertical velocity never recovers. In the NOQR experiment, the maximum vertical velocity actually briefly exceeds that of the control experiment (due to the removal of the rainwater loading within the initial updraft), but then gradually begins to decrease.

The subsequent differences in the maximum vertical velocity evolutions can be better understood by comparing the low-level cold pools generated by the different experiments. Figures 19a-d show the z = 0.25 km perturbation potential temperature at 2305 UTC (26 min into the forecast) for the SDVR control and the three sensitivity experiments. As expected, the SDVR case has the strongest cold pool and the NOQV cold pool is only slightly weaker. For the NOQR case, however, the impact on the cold-pool evolution is more significant. This is consistent with the fact that the NOQR prediction must first generate rainwater before evaporation in the downdraft can generate the low-level cold pool. This weakening of the cold pool leads to a weakening of the low-level convergence and storm updraft. Finally, we see that when neither the rainwater nor the perturbation water vapor are present in the initial fields (NOQQ), only a weak low-level cold pool has formed by 2305 UTC.

Further illustration of the experimental differences is seen in Fig. 20, time series of the three-dimensional rainwater correlation coefficients (with respect to the verification) for the four predictions. By computing the correlation with respect to the verification, we see not only the sensitivity of each experiment relative to the SDVR control, but also the skill of each experiment relative to the dual-Doppler verification. The temporary nature of the perturbation water vapor impact can be seen in the recovery of the NOQV correlation to values very near those of the SDVR control after 26 min. In contrast, the NOQR rainwater field recovers quite rapidly during the first 10 min, but has substantial differences from the SDVR control after 30 min. Finally, the NOQQ correlation time series illustrates the very poor recovery of the rainwater field when neither the perturbation water vapor nor the rainwater is initialized in the model.

Taken as a whole, these moisture sensitivity experiments suggest a strong feedback mechanism from the rainwater field to the cold-pool strength, low-level convergence and updraft strength. Beyond 26 min, the NOQV experiment has very similar correlation coefficients to the SDVR control, while the NOQR experiment actually has higher correlation coefficients (better agreement with the dual-Doppler verification) than the SDVR control. This superiority of the NOQR experiment at later times is consistent with the excessive cold pool noted in the SDVR control, and reinforces the point that the hydrometeor fields exert a strong influence on the subsequent storm evolution.

6. Summary and conclusions

In Part II of this study, we have applied a Gal-Chen (1978) type thermodynamic retrieval procedure to obtain perturbation pressure and potential temperature fields for the three sets of vector wind analyses described in Part I of this study. These wind fields include 1) the dual-Doppler analyzed fields from DB97 (DDOP), 2) the full single-Doppler retrieved winds from Part I (SDVR), and 3) the winds obtained from a simplified retrieval that retained only the observed radial velocity, the estimated mean horizontal wind components, and the perturbation radial divergence contribution to the polar velocity (UVVR). These three sets of wind fields, and the thermodynamic fields obtained from them, are then used to provide initial conditions for short-range numerical predictions of a supercell thunderstorm. Analysis of the prediction accuracy as well as results from simple data sensitivity experiments are then presented.

Prior to the thermodynamic retrieval, a velocity blending and adjustment procedure is applied to the input wind fields. Because the background wind field for this case was limited to a proximity sounding, a fairly simple blending procedure (following Lin et al. 1993) is utilized. It consists of solving a series of 2-D Laplace equations to provide a smooth transition of the horizontal velocity components between the irregular storm edge and the lateral boundaries of the model domain. The variational velocity adjustment retains the observed radial velocity, while enforcing anelastic mass conservation and minimizing departures of the adjusted wind field from the blended fields. The radar-observed reflectivity is converted to rainwater (using a semi-empirical formula) for use in the potential temperature retrieval, while the perturbation water vapor field is specified after the thermodynamic retrieval by saturating updrafts ($w > 3$ m s^{-1}) within the rainwater ($q_r > 0.1$ g kg^{-1}).

The retrieved perturbation pressure fields show good agreement with the linear theory predictions of Rotunno and Klemp (1982). The retrieved perturbation temperature excess in the main storm updraft is in good agreement with expectations from parcel theory calculations using the base-state sounding. The retrieved thermodynamic fields for the simplified wind retrieval are very similar to those for the full SDVR.

Examination of the terms in the vertical momentum equation indicates, as expected, that the main contributions to the retrieved potential temperature are from the vertical pressure gradient, vertical acceleration, and rainwater loading terms. Noting that the ratio of the vertical acceleration to the buoyancy expresses the degree of departure from hydrostatic balance, simple linear theory was used to derive an expression that relates this departure to the aspect ratio of the buoyancy field. Subjective calculation of the relevant quantities for the full single-Doppler retrieved and dual-Doppler fields indicated a reasonable agreement with linear theory. As might be expected, the buoyant updraft in the dual-Doppler case was taller, narrower, and slightly less hydrostatic than its single-Doppler counterpart.

Explicit short-range predictions of the Arcadia supercell thunderstorm are initialized using the three sets of retrieved fields. Qualitative evaluation of the subsequent predictions indicates that for the dual-Doppler derived (DDOP) and full single-Doppler retrieved (SDVR) initial fields, the general storm evolution and deviant storm motion are reasonably well predicted for about 30 min. In contrast, the storm decays quite rapidly in the prediction initialized with the fields obtained from the simplified wind retrieval (UVVR). The similar performance of the dual-Doppler prediction and the full single-Doppler prediction indicates that the single-Doppler velocity retrieval captures those flow features contained in the dual-Doppler analysis that are most important to the storm evolution. Conversely, the poor performance of the simplified retrieval prediction indicates that the portion of the wind field obtained from the pseudo-streamfunction in the full wind retrieval (primarily low-level polar vorticity) is an important factor in accurately predicting the storm evolution (even though it has little impact on the thermodynamic retrieval).

Detailed analysis of the dual-Doppler and full single-Doppler predicted storm evolutions indicate that the main hydrometeor core weakens too rapidly, possibly because the near-surface gust front surges eastward too quickly. This, in turn, may be related to the strength of the low-level cold pool and the initial specification of hydrometeors. This result, combined with the documented sensitivity of the predicted storm evolution to the initial rainwater field, suggests the limitations of neglecting ice-phase microphysics in this study and indicates two areas for further study. First, a more realistic treatment of moisture fields in both the retrieval and the numerical model itself should be investigated. Including ice-phase microphysics in the model predictions is trivial; however, retrieving initial ice-phase fields from the Doppler-radar data is a difficult problem. The availability of polarimetric data would be very helpful in this regard, as shown by Vivekanandan et al. (1999), who have tested a polarimetric microphysical retrieval algorithm. Second, initialization of the thunderstorm induced low-level cold pool is another area that merits further study. The lack of in situ observations combined with the limited ability of operational Doppler radars to observe the boundary layer makes this another difficult problem. Sun and

Crook (1999), however, present encouraging results from the application of their adjoint technique to the problem of boundary layer temperature retrieval using WSR-88D data.

In addition to the sensitivity of the predictions to the initial moisture fields, we have shown the importance of the three-dimensional velocity field relative to the thermodynamic fields. In a set of supercell OSSEs, Weygandt et al. (1999a) found similar results and further demonstrated that their prediction was most sensitive to the perturbation horizontal velocity. This result is somewhat encouraging, given that single-Doppler retrievals likely obtain the horizontal wind with a higher degree of accuracy than the vertical velocity, thermodynamic, or moisture fields. Last, we note that in this study, a simplistic static model initialization was used. Forecast improvement might be obtained by using a cycling technique, in which radar-retrieved fields from successive times are assimilated into the evolving numerical prediction.

This two-part study has documented that a sequential retrieval procedure that combines a single-Doppler velocity retrieval with a thermodynamic retrieval can be used to initialize a realistic short-range numerical prediction of a supercell thunderstorm. Overall, the numerical prediction results suggest at least a degree of short-term predictability for this storm and provide an impetus for continued development of single-Doppler retrieval and assimilation techniques.

Acknowledgements. We are indebted to Howard Bluestein and David Dowell for providing us with the Doppler radar data and dual-Doppler analyses. The radar data were edited and analyzed using software developed at the National Center for Atmospheric Research. Graphics were created using ZXPLOT, developed by Ming Xue. Sue Weygandt assisted with the preparation of figures. The authors have benefited from discussions with Steve Lazarus, Doug Lilly, Jerry Straka, John Lewis, Fred Carr, Scott Ellis and Ming Xue. Scientific reviews by Dezso Devenyi and Steve Koch, and a technical review by Nitta Fullerton are gratefully acknowledged. The research was supported by the National Science Foundation through Grant ATM91-20009 to the Center for Analysis and Prediction of Storms and by a supplemental grant from the FAA. One of us (AS) was also supported by the United States Department of Defense (Office of Naval Research) through Grant N00014-96-1-1112. Computer support was provided by the Environmental

Computing Applications System, which is supported by the University of Oklahoma and National Science Foundation under Grant EAR95-12145.

APPENDIX

Idealized Simulations Using the Proximity Sounding

As an additional benchmark to compare the radar-data initialized numerical predictions against, an idealized "bubble" simulation has been completed using the proximity sounding from this case. This idealized simulation technique has been used for more than 20 years to create numerical storms that exhibit many of the features observed in real convective storms (Schlesinger 1975, Klemp and Wilhemson 1978, Wilhelmson and Klemp 1981, Weisman and Klemp 1984, and many others). More recently, attempts have been made to apply this technique in operational settings to make forecasts of storm types based on observed or model-derived soundings (Brooks et al. 1993, Janish et al. 1995, Wicker et al. 1997).

Using the base-state wind and thermodynamic profiles from the 2115 UTC Tuttle, OK, sounding (Fig. 2), two idealized simulations have been completed on a 100 x 100 km^2 (Δx = 1 km) model domain. In the first, convection was initiated by specifying an ellipsoidal (10 x 10 x 1.5 km^3) region of positive potential temperature excess with a maximum of +2 K in the low levels of the model, while maintaining the relative humidity. Using a moving grid, a 100-min simulation was completed. The convective development was fairly weak and dissipated within 60 min. A second experiment was then conducted, in which a potential temperature excess of 4 K was specified. The initial development was similar to the 2 K experiment; however, redevelopment occurred after 60 min.

Comparison of the domain maximum vertical velocity time series from the 4 K experiment (Fig. B1) to the corresponding time series for the real-data predictions (Fig. 20) indicates some similarities, but also highlights the difficulties of verifying idealized simulations against observed storms. For example, the pronounced decrease in updraft strength beginning at 40 min in the idealized simulation compares favorably with the decrease seen in the real-data predictions and actual storm beginning at 2251 UTC. In reality, however, this decrease in the updraft strength was occurring over 3 h into the lifetime of the storm. Based on the rather short lifetime of the idealized storm, it cannot be classified as a classic supercell. In contrast, the observed Arcadia storm exhibited classic supercell features for much of its greater than 3-h lifetime.

The low-level ($z = 2.25$ km) horizontal vectors and rainwater field at $t = 40, 60, 80$ and 100 min are shown in Figs. B2a-d. Consistent with the vertical velocity maximum time series, the storm appears most intense at $t = 40$ and is weakening by $t = 60$ min. By $t = 80$ min, a new storm has developed along the gust front of the original storm, but it dissipates by $t = 100$ min. The idealized storm from $t = 40$ min does show some resemblance to the actual Arcadia storm during its pretornadic phase; however, the rapid weakening thereafter is not consistent with the observed evolution of the Arcadia storm.

Estimating the potential usefulness of either the idealized simulation or the radar-retrieval based prediction to an operational forecaster is difficult. While the idealized simulation would have suggested the possibility for storms to briefly exhibit supercellular characteristics, it would also have led to an underestimation of the longevity of the storms. The present radar-retrieval based prediction likely also would be of limited use to a forecaster. As a proof-of-concept experiment, however, it does indicate at least some degree of skill for explicit storm-scale predictions initialized with single-Doppler-retrieved fields.

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LIST OF TABLES

LIST OF FIGURES

Figure 4. 2239 UTC N-S cross section (x = 19.5 km) of the perturbation (from the enviromental sounding) winds for the a) DDOP and b) SDVR cases. Contours, grid distances, and radar location are as in Fig. 3.

Figure 5. Vertical profile of maximum three-dimensional anelastic divergence (kg m^{-3} s^{-1}) for the 2239 UTC SDVR retrieved and adjusted wind fields.

Figure 6. Vertical profile of the 2239 UTC retrieved and adjusted maximum vertical velocity (m s^{-1}) for the SDVR case.

Figure 7. 2239 UTC midlevel (z = 5.25 km) a) perturbation pressure and b) vertical vorticity for the SDVR case (shown over a portion of the model domain). Pressure is contoured every 0.2 mb and vorticity is contoured in 10^{-3} s^{-1} increments from $0.1(10^{-3}$ $s^{-1})$. Updrafts greater than 5 m s^{-1} are shaded and the 0.1 g kg^{-1} rainwater contour is depicted as a thick solid line. The mean shear vector (calculated over a 1-km deep layer centered at z = 5.25 km) is indicated by the heavy arrows centered on each of the three updraft maxima.

Figure 8. 2239 UTC N-S cross section (x = 19.5 km) of the retrieved perturbation pressure (contoured in 0.2 mb increments) for the a) DDOP and b) SDVR cases.

Figure 9. 2239 UTC N-S cross section (x = 19.5 km) of a) the retrieved perturbation potential temperature for the SDVR case and the principal contributing terms, including b) the vertical acceleration term, c) the rainwater term and d) the vertical pressure gradient term. Potential temperature and all contributing terms are contoured every 1 K and shown over a portion of the model domain.

Figure 10. Linear theory predicted relationship between the ratio of the vertical acceleration to the buoyancy and the aspect ratio of the buoyancy field. Also plotted are calculated values of the two ratios for the DDOP and SDVR thermodynamic retrievals.

Figure 11. 2239 UTC N-S cross section (x = 19.5 km) of the specified perturbation (from the environmental sounding) water vapor for the SDVR case (shown over a portion of the model domain).

Figure 12. 2251 UTC low-level (z = 2.25 km) rainwater and storm-relative wind vectors for a) the dual-Doppler analysis, and 12-min predictions from b) the DDOP case, c) the SDVR case, and d) the UVVR case. Contours, grid distances, and radar location are as in Fig. 3.

Figure 13. 2305 UTC low-level (z = 2.25 km) rainwater and storm-relative wind vectors (shown over a portion of the model domain) for a) the dual-Doppler analysis, and 26-min predictions from b) the DDOP case, c) the SDVR case, and d) the UVVR case. Contours, grid distances, and radar location are as in Fig. 3.

Figure 14. 2313 UTC low-level (z = 2.25 km) rainwater and storm-relative wind vectors (shown over a portion of the model domain) for a) the dual-Doppler analysis, and 34-min predictions from b) the DDOP case, c) the SDVR case, and d) the UVVR case. ⊗ indicates the location of the main hydrometeor core at the initial time (2239 UTC) and ♦ indicates a persistence forecast calculated from 0-6 km density weighted mean wind. Contours, grid distances, and radar location are as in Fig. 3.

Figure 15. Predicted and observed locations of the z = 3.25 km cyclonic circulation centers at several times. Tracks for a) the various predicted main circulation centers (“M”) and b) the various predicted southern circulation centers (“S”) are shown. The track of

the observed main circulation center is indicated on both panels. Times indicated are UTC.

Table 1. Wind field datasets used for the thermodynamic retrieval and numerical prediction experiments.

NAME	DESCRIPTION
DDOP	**Dual-Doppler analyzed three-dimensional wind fields from Dowell and Bluestein 1997**
SDVR	**Single-Doppler retrieved three-dimensional wind fields (mean wind moving reference frame) from the Cimarron radar**
UVVR	**Simplified three-dimensional wind fields obtained from the Cimarron radial velocity** Observed radial velocity, estimated mean horizontal wind components, and perturbation radial divergence contribution to polar velocity

Table 2. Summary of numerical prediction moisture sensitivity experiments.

NAME	FIELDS REMOVED
SDVR	none
NOQV	q_v'
NOQR	q_r
NOQQ	q_v', q_r